Origin of the high sensitivity of Chinese red clay soils to drought: significance of the clay characteristics

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Abstract

The red clay soils which are widespread in China are known to be highly sensitive to drought during the dry season but the origin of this high sensitivity to drought remains unclear. Several red clay soils were selected in the Hunan province for study. We studied their basic physico-chemical properties and clay mineralogy, their structure and shrinkage properties, as well as their water retention properties. Results show that the amount of water available between –330 and –15 000 hPa water potential is consistent with that recorded in many other clay soils from different parts of the world and thus cannot explain the high sensitivity of the red clay soils to drought. This high sensitivity to drought might be related to the high proportion of poorly available water which was characterized by the amount of available water between –3300 and –15 000 hPa water potential. Comparison with clay soils located in different parts of the world and for which the sensitivity to drought was not identified, showed that this proportion of poorly available water is indeed much higher in the red clay soils studied than in clay soils representing a large range of both clay content and mineralogy. This specific behaviour of the red clay soils studied is thought to be related to the history of their parent materials: these materials are continental sediments which may have been submitted to great hydric stress, thus leading to strongly consolidated soils with consequences such as a high proportion of poorly available water, strong aggregation and weak shrinkage properties.

Keywords: clay mineralogy; porosity; shrinkage; water retention; aggregate; external surface area

1. Introduction

Soils called red soils are widespread in China. They cover 102 million ha, including ten provinces (Cao and Zhu, 1999; He et al., 2004; Wilson et al., 2004a; Zheng et al., 2008). They develop mainly in alluvial Quaternary sediments which reorganize materials resulting from continental alteration or in continental sediments from the Cretaceous or Eocene (BGMRHN, 1988; Wilson et al., 2004a and b; Hu et al., 2010; Shi et al., 2010). The Chinese red soils are either Ferrallisols (Latosolic red soils or Red soil groups) or Semi-Alfisols (Torrid red soil group) of the Genetic Soil Classification of China (Shi et al., 2010). In the International Reference Base System (ISSS Working Group WRB, 2007), Chinese red soils belong to the Alisol, Acrisol or Cambisol group. Most Chinese red soils are Ultisols in the Soil Taxonomy (Soil Survey Staff, 1998), some being Alfisols or Inceptisols (Wilson et al., 2004b). Even when the Chinese red soils have a clay or loamy-clay texture, they are known to be highly sensitive to drought during the dry season (Wang, 1997), thus explaining why the red soil region is called the “red desert of southern China” (Zhao, 2002). The reason for this sensitivity to drought has not yet been elucidated, however. According to Zhang and Zhang (1995), the marked development of micro-aggregation in these red soils may facilitate downward water transfer without enough water storage in the different horizons, thus explaining the poor water storage efficiency of these soils. They also argue that the small amount of water stored in the micro-aggregates is poorly available for biological activity. More recently, Fang et al. (2010) suggested that because of the characteristics of the clays present in the red soils, water retained by these clays is not available enough for biological activity, thus explaining the high sensitivity of the red soils to drought. The clay mineralogy of Chinese red soils has often been described as dominated by kaolinite and oxi-hydroxide minerals (Wilson et al., 2004b). However, several studies have shown a wide variation in the clay mineralogy. Vogel et al. (1995) studied red reference soils from the subtropical Yunnan Province and showed that smectite could be the main clay mineral or present in significant amounts as well as chlorite and illite in addition to kaolinite as the main mineral. Zhang et al. (2004) studied red soils in southern China and showed that the clay mineralogy varied according to the parent material, latitude, elevation and topographic position. Kaolinite and halloysite were the most common clays in these soils, the proportion of kaolinite increasing with the degree of weathering. Vermiculite, illite, chlorite and smectite were also identified as secondary clay minerals. Goethite and hematite were the only Fe oxides occurring in significant quantities in the soils studied by Zhang et al. (2004). This wide variety of clay mineralogy encountered in the Chinese red soils probably indicates that the origin of their high drought sensitivity is not closely related to the clay mineralogy. The objective of the present work is to analyse the water retention properties of a few Chinese red clay soils, to relate these properties to the characteristics of the clays and to compare them to the water retention properties of clay soils from different regions of the world. The results will enable the discussion of the clay characteristics which are responsible for the small amount of available water and hence potentially partially responsible for the high sensitivity of Chinese red clay soils to drought.

2. Materials and methods

2.1 The soils studied

The red clay soils of the Taoyuan experimental station (28°55’47”N, 111°26’33”W) in the Hunan province, 200km west of Changsha were selected for study. This experimental station belongs to the Chinese Ecosystem Research Network. According to the Köppen classification, the climate of the region is Humid Subtropical (Cfa). It is characterized by a dry winter (medium temperature of the coldest month ranging from -3°C to 18°C) and maximum rains in summer. The mean annual rainfall in Taoyuan is 1440 mm, 80% of rainfall occurring between March and August. The mean annual temperature is 16.5°C. The range between winter and summer (difference between the average temperature during winter and summer) is 21°C (Huang et al. 2004). Soils were selected at three locations along a slope (Figure 1): a soil at the top of the slope (TS) under a vegetation mainly composed of tea-oil trees (*Camelia* *oleifera*) and secondarily by camphor laurels (*Cinnamomum camphora*), chestnut trees (*Castanea sp*.) and sandalwood (*Santalum sp.*); a soil at the middle of the slope (MS) in an orchard of orange trees; and a soil on a ledge of the slope (LS) in a cultivated plot after a maize crop, the soil being left bare at the time of soil sampling. The soils are derived from Quaternary red clays. They are Ferric Acrisols in the International Reference Base System (ISSS Working Group WRB, 2007), Udic Ferralosols in the Chinese Soil Taxonomy (CRG-CST, 2001) and Typic Hapludult in the Soil Taxonomy (Soil Survey Staff, 1998). They were sampled in March 2011 after rewetting with 39 and 204 mm of rainfall in February and March, respectively. The soils were thus close to field capacity. A pit 1m in depth was dug and the different horizons were described. Disturbed samples were collected in every horizon as well as undisturbed samples of different volumes.

2.2 Methods

The bulk density (D_b in g cm^{-3}) and field water content of the horizons at sampling date were measured by using cylinders 1236 cm^3 in volume. The bulk density of undisturbed millimetric clods 20 to 30 mm^3 in total volume was measured using the kerosene method (Monnier et al., 1973). It was measured in triplicate with 5 to 10 millimetric clods at a water content corresponding to sampling conditions and after air-drying in the laboratory (Bruand and Prost, 1987). The particle size distribution was measured using the pipette method after pre-treatment of samples with hydrogen peroxide and sodium hexametaphosphate (Robert & Tessier, 1974). The cation exchange capacity (CEC, in cmol$_+$ kg^{-1} of oven-dried soil) and the exchangeable cations were measured using the cobalt-hexamine trichloride method (Ciesielski & Sterckeman 1997) and organic carbon content (*OC*) by oxidation using excess potassium bichromate in sulphuric acid at 135°C (Baize, 2000). The gravimetric water content was determined at –60, –100, –330, –1000, and –3300 hPa water potential by using in triplicate soil cores 100 cm^3 in volume and at –15000 hPa water potential by using undisturbed clods (10–15 cm^3 in volume) collected when the soil was near to field capacity (Bruand and Tessier, 2000). The soil cores were thoroughly saturated for one week before they were placed inside a pressure plate chamber (Soil Moisture Equipment Corp, Santa Barbara, USA) to drain at the sequence of pressure from –60 to –3300 hPa (Jing et al., 2008). The mineralogical composition of the <2µm material of the horizons B1 and C of each soil was determined by X-ray diffraction (XRD). Samples were air-dried and manually ground to a powder in a mortar. The clay fraction was collected using the sedimentation method at 20°C after mechanical dispersion. Oriented clay deposits were prepared on glass slides and studied using an X-ray diffractometer equipped with a Si(Li) solid detector to filter the $Cu_{K\alpha}$ radiation ($\lambda Cu_{K\alpha}$ = 1,5418 Å) of a standard European type X-ray tube (40kV, 40mA). The divergence, the incident beam scatter, the diffracted beam scatter and the receiving slits were 2.00, 4.00, 0.50 and 0.22 mm wide, respectively. The XRD patterns were collected from 2° to 24° 2θ at a scan rate of 0.3° 2θ/min per step of 0.05° 2θ, on natural, glycoled and heated oriented slides (Robert and Tessier, 1974; Bruand and Prost, 1988). Nitrogen adsorption isotherms were conducted on about 150 mg of < 2 mm soil samples dried in a 105ºC oven for 24 hours and then dried again at 105°C under a pressure of 10^2 Pa. The specific surface area (*SSA*) of the material was determined using the BET equation (Brunauer et al., 1938). Experiments were performed with a surface analyser (Model Nova 2200e, Quantachrome Instrument Company, USA).

3. Results and discussion

3.1 Basic physico-chemical characteristics

The clay content (*CC* in g per kg of oven-dried soil at 105°C) increased roughly with depth in each soil and ranged from 373 to 474 g kg^{-1} in soil TS, from 359 to 461 g kg^{-1} in soil MS and from 255 to 378 g kg^{-1} in soil LS (Table 1). These clay contents are similar to those reported for red soils by Baligar et al. (2004). The fine silt (2-20 µm) content ranged from 382 to 505 g kg^{-1} (Table 1). A high fine silt content was also recorded in other red soils located in southern China (Hong et al., 2010). The pH ranged from 4.2 to 4.7 and increased with depth in the three soils studied (Table 1). These values are consistent with those earlier recorded in southern China for red soils under upland tree vegetation and crops (Xu et al., 2003). The cation exchange capacity (*CEC* in $cmol_+$ per g of oven-dried soil) was very small considering the clay content and poorly saturated by the alkaline cations (about 10-20 %); exchangeable aluminium (not measured here) was the main exchangeable cation, as pointed out by Xu et al. (2003).

Because the subsoil horizons contained little organic carbon (Table 1), we assumed that the contribution of the organic matter to the cation exchange capacity was negligible compared with the cation exchange capacity of the clay. Thus we calculated for the subsoil horizons the cation exchange capacity of the clay (CEC_{cl}, in $cmol_+$ per g of oven-dried clay at 105°C) as follows (Bruand and Tessier, 2000):

$$CEC_{cl} = (CEC / CC) \times 1000$$

Results showed that CEC_{cl} ranged from 9.58 to 14.11 $cmol_+ g^{-1}$ which corresponded to a clay fraction mainly made up of non active clays (Table 1). XRD patterns of the clay fraction extracted from the horizons C show the presence of:

- 1:1 clays (d001 = 0.713 nm) which are in all likelihood kaolinite, as the peak at 0.713 nm is similar at 20°C, 170°C and after treatment with ethylene glycol (Figure 2). The much smaller peak at 0.713 nm recorded after heating at 500°C is thought to be related to the remaining presence of kaolinite which is not dehydroxylated at such a temperature (Robert and Tessier, 1974; Bruand and Prost, 1986);
- non swelling 2:1 clays (d001 = 1,004 nm) which are in all likelihood particles of micas < 2 µm or large particles of illite because of the very small CEC_{cl} recorded, the peak at 1,004 nm being similar at 20°C, 170°C and after treatment with ethylene glycol (Figure 2) (Robert and Tessier, 1974);
- and non swelling 2:1/1 clays (d001 = 1.400 nm) which are probably chlorites or hydroxy-vermiculites, the peak at 1.400 nm being similar at 20°C and after treatment with ethylene glycol (Figure 2) (Robert and Tessier, 1974). At 170°C, the X-ray diagrams show a decrease in the intensity of the peak at 1.400 nm probably because of the partial destruction of the octahedra in the interlamellar space. At 500°C, the peak at 1.400 nm was either very small (Figures 2c and f) or absent (Figures 2a, b, d and e), thus indicating the nearly total or total destruction of the octahedra in the interlamellar space.

The clay mineralogy appears therefore to be characterized by the presence of non swelling or poorly swelling clays. A similar clay composition with some illite/smectite interstratified clays was described by Hong et al. (2010) in red soils located in southern China. Because of the lack of a swelling test using glycol or ethylene glycol in their study, the illite/smectite interstratified clays identified by the presence of a peak at 1.464 nm might in fact be chlorites or hydroxy-vermiculites such as for the soils selected for this study (Robert and Tessier, 1974).

The specific surface area (*SSA* in m^2 per g of oven-dried soil) ranged from 9.9 to 34.1 $m^2 g^{-1}$ and the clay content accounts for 84 % of its variance. As contributions to *SSA* from silt and sand are negligible for clay soils, the major contribution is likely to be made by clay-size materials. We thus calculated the *SSA* of the clay fraction (SSA_{cl}, in m^2 per g of oven-dried clay) for horizons B1, B2 and C as follows:

$$SSA_{cl} = (SSA / CC) \times 1000.$$

The results showed that SSA_{cl} was between 41.5 and 77.3 $m^2\ g^{-1}$ (Table 1) thus indicating a small size of the <2 µm fraction in these horizons (e.g. Feller et al., 1992; De Brito Galvão and Schulze, 1996; Balbino et al., 2002). Reatto et al. (2009) studied Brazilian clayey Acrisols (ISS Working Group WRB, 2007) and found that SSA_{cl} was between 24.1 and 41.1 $m^2\ g^{-1}$ for a clay fraction made of kaolinite, gibbsite, goethite and hematite particles between 0.01 and 0.3 µm. Bruand and Tessier (2000) discussed the water retention properties of French clayey soils developed on a large range of clayey sediments and found that SSA_{cl} was between 50 and 130 $m^2\ g^{-1}$ for a clay fraction made of clay particles with a small number of layers (Robert et al., 1991).

3.2 Structure and porosity

The red soils studied show essentially a moderate to strong subangular blocky structure 5 to 10 mm in size (Table 2), thus making it easy to separate millimetric aggregates as defined by Bruand and Prost (1987). Such a strong structure is consistent with the characteristics described by Zhang and Zhang (1995) for red soils. Roots were mainly located in the A and B1 horizons in TS and MS where the soils were under a secondary forest vegetation and an orchard of orange trees, respectively. In soil LS, roots remaining from the maize crop were not observed.

In each soil, the bulk density was lower in the horizons A compared to its value in the subsoil horizons because of the lesser development of biological pores in the latter (Table 3). It ranged from 1.27 to 1.32 g cm^3 in soil TS, from 1.34 to 1.42 g cm^3 in soil MS and from 1.45 to 1.46 g cm^3 in soil LS. These values are close to those reported by Vogel et al. (1995) for red reference clay soils of the Yunnan Province. The bulk densities recorded for the top horizon in soils TS, MS and LS are similar to those recorded for many topsoils of red soils under different soil managements in southern China (e.g. Huang et al., 2004; Zheng et al., 2008; Huang et al., 2010) but the clay content was often less than 30 % in the soils studied.

These bulk densities are also close to those recorded by Bruand and Tessier (2000) for French clay soils with both a large clay content and mineralogy range.

The volume of macropores and micropores in the soil at the sampling date (i.e. close to field capacity), was computed by using the bulk density of the horizon and the bulk density of the millimetric aggregates (Bruand and Prost, 1987). Thus, the specific volume of macropores in conditions close to field capacity (V_M^{FC}) was obtained by subtracting the specific volume of the millimetric aggregates (reciprocal of their bulk density) from the specific volume of the horizons (reciprocal of their bulk density) (Table 3). The results showed that V_M^{FC} decreases strongly with depth in the three soils studied (Table 3). Such results are consistent with the few or very few roots recorded in horizons B1, B2 and C of these soils (Table 2). The volume of micropores at field capacity V_m^{FC} and after air drying V_m^{AD} was computed by subtracting the specific volume of the solid phase which is equal to the reciprocal of the particle density, the latter being considered as equal to 2.65 g cm^{-3} (1/2.65 g cm^{-3} = 0.377 cm^3 g^{-1}) (Bruand and Prost, 1987), from the specific volume of the millimetric aggregates (Table 2). The results showed that V_m^{FC} is smaller in the subsoil (horizons B1, B2 and C) than in horizon A for each soil. In the subsoil, V_m^{FC} ranges from 0.239 to 0.297 cm^3 g^{-1} and V_m^{AD} from 0.208 to 0.264 cm^3 g^{-1}. In clay soils the volumes of micropores result mainly from the assemblages of the clay particles in the soil material (Fiès and Bruand, 1990; 1998). We therefore computed the volume of micropores developed by the clay phase close to field capacity ($V_{m_{clay}}^{FC}$ in cm^3 per g of clay) and after air-drying ($V_{m_{clay}}^{AD}$ in cm^3 per g of clay) as follows:

$$V_{m_{clay}}^{FC} = (V_m^{FC}/CC) \times 1000,$$

$$V_{m_{clay}}^{AD} = (V_m^{AD}/CC) \times 1000.$$

The average $V_{m_{clay}}^{FC}$ and $V_{m_{clay}}^{AD}$ computed with the data presented in Tables 1 and 3 for horizons B and C with *CC*>300 g kg^{-1} are 0.663 and 0.584 cm^3 per g of clay, whereas the corresponding values for a horizon B from a French red clay soil are 0.488 and 0.327 cm^3 per g of clay, respectively (Bruand and Prost, 1987). This corresponds to an 11% decrease in the average micropore volume of the clay phase between field capacity and air-drying for the Chinese red clay soils studied, compared to 33% for the clay soil studied by Bruand and Prost (1987). Thus the clay fabric which corresponds to both the volume of clay particles and associated pore volume resulting from their assemblage shrinks much less between field capacity and air-drying in the Chinese red clay soils studied than in the French clay soil studied by Bruand and Prost (1987).

3.3 Water retention properties

The amounts of water retained at –60, –100, –330, –1000, –3300 and –15 000 hPa water potential for the different horizons studied are given in Table 4. The water content at the sampling date corresponded to a water potential between –100 and –330 hPa, thus confirming that the soil was close to field capacity during sampling (Al Majou et al., 2008a). The variance in water retained at –60, –100, –330 hPa water potential is explained mainly by the specific volume of the horizon (reciprocal of the bulk density) with R^2 = 0.863, 0.826 and 0.646, respectively. At –15 000 hPa water potential, the maximum variance in water retained is explained by the clay content (R^2 = 0.851), while the specific volume of the horizon explained only a small proportion of the variance. Such a high proportion of variance explained by the specific volume and the clay content of the horizon at high and low water potential, respectively, was earlier recorded by Bruand et al. (1988) and Bruand (1990) for clayey soils.

For the horizon with a clay content > 300 g kg^{-1}, we computed the available water which corresponds to the amount of water available between –330 and –15 000 hPa water potential (Table 5). It ranged from 0.0091to 0.166 cm^3 cm^{-3} in the soils studied here. These values are consistent with those recorded in many other clay horizons originating from soils located in different countries (Table 5). The average available water measured for the Chinese red clay soils of this study is however slightly greater (0.121 cm^3 cm^{-3}) than for the 1203 clayey soils from the literature (0.113 cm^3 cm^{-3}) (Table 5). The available water between –3300 and –15 000 hPa reported in the literature for Chinese red clay soils is much greater (average of 0.222 cm^3 cm^{-3}) but the soils selected were mainly topsoils (Table 5).

Although the available water between –3300 and –15 000 hPa of the Chinese red clay soils studied appears to be slightly greater than clay soils from other countries, a much greater proportion of water available between –3300 and –15 000 hPa compared to the water available between –330 and –15 000 hPa was recorded for these Chinese red clay soils than for clay soils originating from other countries (Table 5). This proportion ranged from 0.520 to

0.753 in the Chinese red clay soils of this study (average of 0.620), from 0.535 to 0.620 in the Chinese red clay soils from the literature (average of 0.591), and from 0.285 to 0.425 in the 1203 clay soils from other countries (average of 0.366) (Table 5). This indicates that for a given available water amount, the proportion of water weakly available is much greater in Chinese red clay soils than in clay soils originating from other countries, the total available water being just slightly greater for the former. This might explain the high sensitivity to drought of the Chinese red clay soils during the dry season as pointed out by Wang (1997) and Zhao (2002).

In most clay soils, a proportion of water is released at low water potential by the clay phase concomitantly to its shrinkage (Bruand and Prost, 1987). The millimetric aggregates originating from the red clay soils studied showed little shrinkage between field capacity and air-drying (Table 2). Such a small shrinkage which reflects a stiffness of the microstructure is likely related to the high proportion of non active clays as revealed by the cation exchange capacity and the X-ray diffraction patterns (Table 1 and Figure 2) and the presence of aluminum and iron oxy-hydroxides as revealed by the soil color but not studied here. This small shrinkage and the high proportion of poorly available water might also be related to the long hydric stress history of the Chinese red clay soils as discussed by Bruand and Tessier (2000) for a large set of French clayey soils developed on clayey sediments. According to their results and those recorded earlier by Tessier and Pédro (1987) on pure clays, the water retention properties of clayey soils appear to be mainly affected by the greatest effective stress recorded by the soil, the proportion of poorly available water compared to the total amount of available water increasing with the effective stress recorded. The parent materials of the Chinese red clay are mainly Quaternary continental sediments which reorganize material resulting from continental alteration from the Cretaceous or Eocene (BGMRHN, 1988; Wilson et al., 2004a and b; Hu et al., 2010; Shi et al., 2010). Because of this history, they were in all likelihood submitted to great hydric stresses, thus leading to strongly consolidated clayey parent materials with consequences for the soils such as a high proportion of poorly available water, strong aggregation, and weak shrinkage properties (Bruand and Tessier, 2000).

Conclusion

Our results show that the amount of water available between –330 and –15 000 hPa water potential for the Chinese red clay soils selected in the Taoyuan experimental station is consistent with that recorded in many other clay soils and thus cannot explain the high sensitivity of the Chinese red clay soils to drought. Although many soil and plant characteristics can potentially be involved in the high sensitivity of Chinese red clay soils to drought, the high proportion of poorly available water which was characterized by the amount of water available between –3300 and –15 000 hPa water potential compared to the available water between –330 and –15 000 hPa might explain this sensitivity to drought. Comparison with clay soils located in different regions of the world and for which a particular sensitivity to drought was not identified, showed that this proportion of poorly available water is indeed much greater in the Chinese red clay soils studied. Finally, our results which were recorded on a small number of soils need to be confirmed by studies on a much larger range of Chinese red clay soils coming from different areas. By providing more accurate knowledge of parent and soil history, this would enable discussion of the amount of water available between –3300 and –15 000 hPa water potential.

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Table 1

Physico-chemical characteristics of the soils studied at the top (TS), the middle (MS) and on a ledge of the slope (LS)

Soil	Horizon	Particle size distribution (μm)					Organic carbon	pH	Cation exchange capacity (*CEC*) and exchangeable cations					CEC_{cl}	Specific surface area (*SSA*)	SSA_{cl}
		<2	2 – 20	20 – 50	50–200	200 – 2000			*CEC*	Ca^{2+}	Mg^{2+}	Na^{+}	K^{+}			
		g kg^{-1}					g kg^{-1}		$cmol_{+}$ kg^{-1}					$cmol_{+}$ kg^{-1}	m^2 kg^{-1}	m^2 kg^{-1}
TS	A	373	470	109	21	27	35.40	4.3	6.37	0.71	0.24	0.18	0.02	-	16.0	-
	B1	441	433	93	12	21	5.44	4.4	4.52	0.41	0.08	0.17	0.02	10.25	26.2	59.4
	B2	497	382	85	12	24	2.13	4.5	6.56	0.32	0.08	0.09	0.03	13.20	36.8	74.0
	C	474	398	86	16	26	2.05	4.7	6.69	0.42	0.12	0.09	0.03	14.11	34.7	73.2
MS	A	359	493	111	13	24	17.51	4.2	4.36	0.18	0.10	0.16	0.10	-	11.8	-
	B1	380	478	107	12	23	4.4	4.3	3.64	0.12	0.06	0.09	0.01	9.58	24.1	63.4
	B2	441	433	92	11	23	2.73	4.7	5.55	0.46	0.18	0.09	0.02	12.59	34.1	77.3
	C	461	413	92	11	23	2.44	4.7	6.34	0.45	0.15	0.10	0.02	13.75	32.3	70.1
LS	A	255	490	142	37	76	15.23	4.4	3.68	1.55	0.39	0.14	0.02	-	9.9	-
	B1	237	505	151	35	72	7.63	4.3	3.16	0.43	0.21	0.11	0.01	13.33	8.8	37.1
	B2	322	485	118	21	54	5.24	4.6	3.30	1.10	0.18	0.08	0.02	10.25	16.6	51.6
	C	378	449	99	25	49	6.15	4.4	4.35	0.79	0.16	0.11	0.02	11.51	15.7	41.5

Table 2

Morphological characteristics of the horizons in soil TS, soil MS and soil LS

Pits	Horizons	Depth	Matrix Munsell color		Structure			Root density[(2)]
		(cm)	Wet	Dry	Type	Grade[(1)]	Size (mm)	
TS	A	3-10	7.5YR 3/3	10YR 5/4	Granular	3	1-5	4
	B1	10-40	5YR 5/6	7.5YR 6/6	Subangular blocky	3	5-10	3
	B2	40-80	5YR 4/6	5YR 5/6	Subangular blocky	3	5-20	1
	C	80-100	5YR 4/6	5YR 5/8	Subangular to angular blocky	2	5-10	1
MS	A	0-10	5YR 4/4	5YR 7/3	Granular to subangular blocky	3	1-5	3
	B1	10-40	5YR 4/4	5YR 7/4	Subangular blocky	3	5-10	2
	B2	40-80	5YR 4/6	5YR 7/4	Subangular blocky	3	5-15	1
	C	80-100	5YR 3/6	5YR 7/4	Subangular blocky	2	5-20	1
LS	A	0-15	7.5YR 4/6	10YR 5/4	Subangular blocky	2	5-15	0
	B1	15-35	7.5YR 5/6	7.5YR 6/6	Subangular blocky	2	5-30	0
	B2	35-65	7.5YR 5/6	7.5YR 6/6	Subangular blocky	2	5-30	0
	C	65-90	7.5YR 5/6	7.5YR 6/6	Subangular blocky	1	5-30	0

[(1)]Grade: 1=weak; 2=moderate; 3=strong. [(2)]Root frequency: 0=very few (<1 root in each horizontal 100cm^2 surface area); 1= few (<1 to 5 roots in each horizontal 100cm^2 surface area); 2 = moderate (5 to 10 roots in each horizontal 100cm^2 surface area); 3=numerous (10 to 50 roots in each horizontal 100cm^2 surface area); 4=highly numerous (>50 roots in each horizontal 100cm^2 surface area).

Table 3
Bulk density of the horizons (cylinder method) at sampling date and of millimetric aggregates (kerosene method) at sampling date and after air-drying (standard deviation between brackets)

Soil	Horizon	Bulk density at sampling date: Horizon(a)	Bulk density at sampling date: Millimetric Aggregates(b)	Bulk density after air-drying: Millimetric Aggregates(c)	Volume of pores at sampling date: Macropores (1/a)-(1/b)	Volume of pores at sampling date: Micropores (1/b)-0.377	Volume of micropores in the dried soil material (1/c)-0.377
		$g\ cm^{-3}$		$g\ cm^{-3}$	$cm^3\ g^{-1}$		$cm^3\ g^{-1}$
TS	A	1.04 (0.04)	1.37 (0.06)	1.49 (0.04)	0.232 (0.018)	0.353 (0.022)	0.294
	B1	1.32 (0.06)	1.51 (0.01)	1.56 (0.02)	0.094 (0.006)	0.287 (0.004)	0.264
	B2	1.39 (0.03)	1.57 (0.06)	1.67 (0.01)	0.082 (0.014)	0.260 (0.023)	0.222
	C	1.44 (0.03)	1.62 (0.01)	1.71 (0.02)	0.078 (0.005)	0.239 (0.002)	0.208
MS	A	1.07 (0.01)	1.44 (0.02)	1.52 (0.03)	0.240 (0.016)	0.318 (0.011)	0.281
	B1	1.34 (0.07)	1.52 (0.01)	1.59 (0.01)	0.087 (0.008)	0.282 (0.001)	0.252
	B2	1.39 (0.03)	1.49 (0.05)	1.59 (0.02)	0.045 (0.014)	0.297 (0.023)	0.252
	C	1.42 (0.04)	1.54 (0.03)	1.60 (0.02)	0.053 (0.016)	0.274 (0.012)	0.248
LS	A	1.21 (0.03)	1.45 (0.03)	1.52 (0.03)	0.137 (0.018)	0.313 (0.016)	0.281
	B1	1.45 (0.09)	1.62 (0.02)	1.66 (0.03)	0.074 (0.007)	0.239 (0.007)	0.225
	B2	1.46 (0.01)	1.53 (0.02)	1.61 (0.04)	0.033 (0.012)	0.275 (0.010)	0.244
	C	1.45 (0.01)	1.50 (0.04)	1.60 (0.02)	0.024 (0.018)	0.289 (0.016)	0.248

Table 4
Water content at sampling date and water retention properties of the soils studied.

Soil	Horizon	Water content at sampling date	Water retained at a water potential of (hPa) :					
			–60	–100	–330	–1000	–3300	–15 000
		g g^{-1}						
TS	A	0.345	0.394	0.375	0.340	0.315	0.286	0.210
	B1	0.273	0.289	0.278	0.252	0.234	0.215	0.175
	B2	0.295	0.318	0.307	0.288	0.275	0.255	0.220
	C	0.286	0.305	0.297	0.284	0.272	0.253	0.212
MS	A	0.315	0.378	0.355	0.310	0.275	0.246	0.155
	B1	0.265	0.281	0.275	0.261	0.249	0.234	0.172
	B2	0.284	0.299	0.296	0.283	0.272	0.256	0.213
	C	0.289	0.299	0.297	0.287	0.279	0.264	0.223
LS	A	0.330	0.341	0.331	0.303	0.278	0.247	0.111
	B1	0.258	0.265	0.261	0.246	0.233	0.212	0.106
	B2	0.259	0.269	0.266	0.255	0.242	0.227	0.141
	C	0.270	0.279	0.276	0.266	0.254	0.240	0.176

Table 5

Available water between –330 hPa (θ_{330}) and –15000 hPa (θ_{15000}) and proportion of this available water between –3300 hPa (θ_{3300}) and –15000 hPa (θ_{15000})

Origin of soils	Reference	Number of horizons	Type of horizon	Clay content	θ_{330}	θ_{3300}	θ_{15000}	$(\theta_{330} - \theta_{15000})$	$(\theta_{3300} - \theta_{15000})$ / $(\theta_{330} - \theta_{15000})$
				%	cm^3 cm^3				
				This study					
TS	This study	1	Topsoil (A)	37	0.354	0.297	0.218	0.136	0.588
		1	Subsoil (B1)	44	0.333	0.284	0.231	0.102	0.520
		1	Subsoil (B2)	50	0.400	0.355	0.306	0.094	0.521
		1	Subsoil (C)	47	0.409	0.364	0.305	0.104	0.567
MS	This study	1	Topsoil (A)	36	0.332	0.263	0.166	0.166	0.584
		1	Subsoil (B1)	38	0.350	0.314	0.230	0.120	0.700
		1	Subsoil (B2)	44	0.393	0.356	0.296	0.097	0.619
		1	Subsoil (C)	46	0.409	0.375	0.317	0.091	0.637
LS	This study	1	Subsoil (B2)	32	0.372	0.331	0.206	0.166	0.753
		1	Subsoil (C)	38	0.386	0.348	0.255	0.131	0.710
Average (this study)								0.121	0.620
				Chinese red clay soils					
China	Cao and Zhu (1999)	1	n.d.	n.d.[a]	0.251	0.198	0.137	0.114	0.535
	Lu et al. (2004)[b]	1	Topsoil	n.d.[c]	0.443	0.340	0.172	0.271	0.620
		1	Topsoil	n.d.[d]	0.447	0.331	0.151	0.296	0.608
		1	Topsoil	n.d.[e]	0.372	0.289	0.165	0.207	0.600
Average (Chinese red clay soils)								0.222	0.591
				Clay soils from literature					
Belgium	Vereecken et al. (1989)	3	n.d.	>35	0.439	0.340	0.293	0.109	0.285

Brazil	Reatto et al. (2007)	1	Subsoil	52	0.267	0.228	0.205	0.062	0.365
		1	Subsoil	61	0.263	0.226	0.208	0.049	0.328
		1	Subsoil	75	0.265	0.232	0.216	0.049	0.339
		1	Subsoil	55	0.277	0.233	0.206	0.071	0.371
		1	Subsoil	78	0.287	0.256	0.239	0.048	0.352
China	Qu et al. (2009)	1	Subsoil	32	0.338	0.244	0.202	0.136	0.308
		1	Subsoil	32	0.429	0.372	0.343	0.087	0.343
		1	Subsoil	32	0.356	0.297	0.268	0.087	0.325
Europe	Wösten et al. (1999)	21	Topsoil	>60	0.489	0.390	0.336	0.153	0.358
		217	Topsoil	35 - 60	0.404	0.324	0.279	0.125	0.357
		596	Subsoil	>60	0.407	0.338	0.298	0.108	0.368
		132	Subsoil	35 - 60	0.471	0.404	0.363	0.109	0.378
France	Al Majou et al. (2008b)	2	Topsoil	>60	0.402	0.332	0.293	0.109	0.358
		17	Topsoil	35 - 60	0.367	0.304	0.272	0.095	0.337
		18	Subsoil	>60	0.405	0.354	0.330	0.075	0.320
		85	Subsoil	35 - 60	0.348	0.298	0.261	0.087	0.425
Portugal	Goncalves et al. (1997)	18	Topsoil and subsoil	>40	0.372	0.302	0.263	0.110	0.359
		38	Topsoil and subsoil	40 - 60	0.426	0.322	0.269	0.158	0.335
USA	Schaap and Leij (1998)	34	Topsoil and subsoil	>40	0.320	0.206	0.160	0.161	0.287
		12	Topsoil and subsoil	35 - 55	0.376	0.241	0.181	0.195	0.306
	Pucket et al. (1985)	1	Subsoil	33	0.321	0.287	0.272	0.049	0.316
		1	Subsoil	34	0.351	0.325	0.313	0.038	0.308
		1	Subsoil	35	0.356	0.319	0.297	0.060	0.368
Weighted average (clay soils from the literature)								0.113	0.366

[a]: soil derived from a Quaternary red clay with a clay content ≥ 0.362 g kg^{-1}; [b]: estimated from soils I.1, I.2 and I.3 in figure 1; [c]: silty clay; [d]: clay; [e]: silty clay;

Figure 1
Location of the soils studied in the Taoyuan experimental station (soil at the top of slope: TS, middle of the slope: MS and on a ledge of the slope: LS)

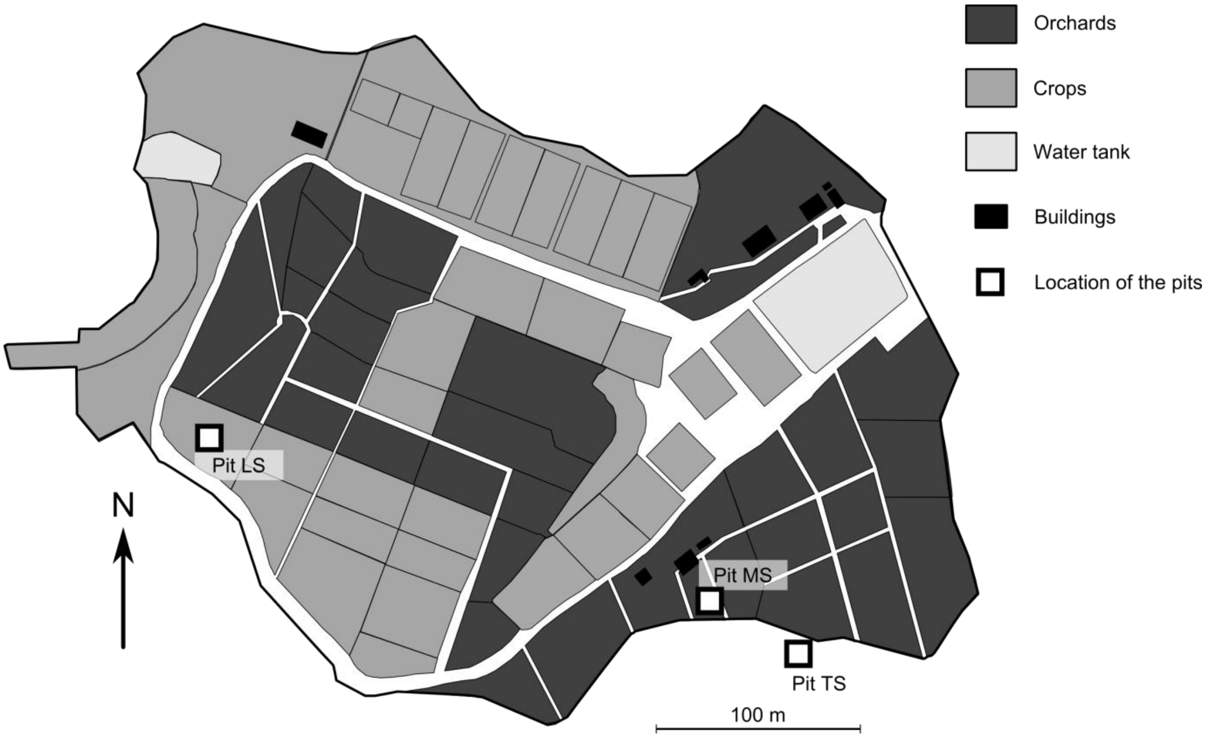

Figure 2

X-ray diagrams of oriented deposits of the <2µm material extracted from horizon B1 of the soil TS (a), MS (b) and LS (c) and from horizon C of the soil TS (d), MS (e) and LS (f) under room conditions (20°C), after saturation with ethylene glycol (EG), after heating at 170°C and 500°C